An organic geochemical record of Sierra Nevada climate since the LGM from Swamp Lake, Yosemite

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1. Introduction

The pacing of natural climatic variability is of particular importance in California, where a limited water supply is dependent on snowmelt runoff from the Sierra Nevada, and is thus sensitive to changes in mean values and seasonal patterns of precipitation and temperature (Wilkinson et al., 2002; Miller et al., 2003). In light of the short instrumental record, only paleoclimatic records can reveal patterns of variability operating on timescales longer than a few decades. These patterns may influence regional water supplies and the occurrence of extreme events such as floods and droughts, and obscure or intensify the effects of human interference with global climate.

Climate variability in California and western North America is affected by Pacific Ocean surface conditions that influence basin-wide atmospheric circulation patterns (Seager and Vecchi, 2010). For example, interannual cycles in tropical Pacific sea surface temperature (SST) (El Niño-Southern Oscillation, ENSO) are associated with anomalous temperature and precipitation throughout western North America and warm SST off California (e.g., Redmond and Koch, 1991; McGowan et al., 1998; Dettinger et al., 2001; Diaz et al., 2001; Higgins et al., 2002; Castello and Shelton, 2004). Likewise, decadal shifts in North Pacific SST and atmospheric circulation are linked to changes in regional temperature, precipitation, snow accumulation, streamflow and runoff timing, and related ecological processes (Dettinger and Cayan, 1995; Mantua et al., 1997; McCabe and Dettinger, 2002; Millar et al., 2004; Tootle et al., 2005; Hunter et al., 2006). The regional influences of...
ENSO and the Pacific Decadal Oscillation (PDO) have been detected in paleorecords spanning the last millennium (e.g., Biondi et al., 2001; Benson et al., 2003; Cook et al., 2004; MacDonald and Case, 2005; Graham et al., 2007; Herweijer et al., 2007; Conroy et al., 2009), but our understanding of the relationship between ocean conditions and climate in western North America over longer timescales, and during climate regimes of the more distant past, remains fragmentary (Barron and Anderson, 2011). This is particularly true of the large Sacramento-San Joaquin River watershed draining the Sierra Nevada.

To date, only a handful of published paleoclimatic records from sites in the Sierra Nevada region combine high temporal resolution (centennial-scale and finer) and length (full Holocene or longer) (e.g., Hughes and Graumlich, 1996; Benson et al., 2002; Mensing et al., 2004). Previous research has linked millennial-scale climatic and environmental change to insolation, continental ice sheet dynamics and local glacial advances/retreats, and various ocean processes (e.g., Davis et al., 1985; Davis and Moratto, 1988; Anderson, 1990; Smith and Anderson, 1992; Anderson and Smith, 1994; Koehler and Anderson, 1994, 1995; Benson et al., 1996, 1997, 1998; Davis, 1999a; b; Mensing, 2001; Brunelle and Anderson, 2003; Porinchu et al., 2003; Mensing et al., 2004; Bacon et al., 2006; Negri et al., 2006; Oster et al., 2009). However, these studies have only occasionally addressed variability on sub-millennial timescales during climate regimes prior to the late Holocene.

In the present study we use organic geochemical analyses of bulk sediments from Swamp Lake (SL), a mid-elevation lake in Yosemite National Park (Fig. 1), to reconstruct paleoenvironmental variability in the Sierra Nevada at ~100-year resolution over the past ~20,000 years. Lake sedimentary organic matter (OM) derives from both autochthonous production (algae, bacteria and aquatic plants) and the input of terrestrial plant material from the surrounding watershed, and is thus an important archive of paleoenvironmental change in the lake basin. Carbon and nitrogen elemental abundances (TOC, TN, C/N) and isotopic compositions (δ¹³Corg, δ¹⁵N) in sedimentary OM respond to changes in the balance of OM sources, vegetation distribution, primary production and lake trophic status, nutrient inputs and sources, and carbon utilization and cycling (e.g., Meyers and Takemura, 1997; Hodell and Schelske, 1998; Brenner et al., 1999; Meyers, 2003). These proxies may therefore provide sensitive indicators of climate-driven environmental change (Meyers and Takemura, 1997; Meyers and Lalier-Verges, 1999; Talbot and Laerdal, 2000). At SL, we apply OM proxies to gather new insight into the Late Pleistocene–Holocene climate history of the Sierra Nevada. In particular, we attempt to better define the major climatic regimes of the last 20 kyr, patterns of century-scale variability within these regimes, and their relationship to existing reconstructions of SST in the eastern North Pacific (e.g., Seki et al., 2002; Barron et al., 2003). The SL record provides a western Sierra point of comparison with records from eastern Sierra drainages (Benson et al., 1996, 1997, 1998, 2002, 2010; Davis, 1999a; Mensing et al., 2004; Briggs et al., 2005; Bacon et al., 2006), and contributes to the synthesis of a more complete picture of regional paleoclimate in western North America since the last glacial maximum (LGM).

2. Materials and methods

2.1. Study area

Swamp Lake is a mid-elevation tarn lake located in Yosemite NP in the central Sierra Nevada (1554 m elevation, 37°57′N, 119°49′W (Fig. 1). Thanks to its location less than 2 km east of the maximum westward (and downslope) edge of Tioga stage glaciation (Smith and Anderson, 1992), SL was ice-free earlier than high-elevation sites, and the sediments accumulating here provide the longest continuous paleoenvironmental record yet recovered from the western Sierra. The granitic SL basin covers ~1.3 km², while the lake itself has a surface area of 0.08 km² and reaches a maximum depth of 20 m. Inlet and outlet streams to the lake are ephemeral, operating only from the winter-wet season to early summer, though perennial subsurface inflow and outflow may occur. At present, SL does not freeze during the winter, with water temperatures reaching a minimum of ~5 °C (Fig. 2a). Continuous air temperature readings taken in Dec–Feb of 2006–2007 fluctuated between ~7 and 5 °C (Roach, 2010), with no below-zero excursion.
lasting long enough to induce lake freezing. Temperature stratification in SL develops over the course of the spring–summer growing season (Fig. 2a) following deep mixing/isothermal conditions during the winter. Chlorophyll a (chl a) profiles taken between March and August (2008–2010) reveal a deep (10–15 m) chl a maximum lasting through the summer, following an early spring (Mar–Apr) bloom in the epilimnion (Fig. 2b).

Terrestrial vegetation at the site is of the Sierra lower montane forest type. Common species include ponderosa pine (Pinus ponderosa), sugar pine (Pinus lambertiana), incense cedar (Calocedrus decurrens), white fir (Abies concolor), manzanita (Arctostaphylos spp.) and black oak (Quercus kelloggii). Riparian species along the lake margin include western azalea (Rhododendron occidentale), mountain alder (Alnus rhombifolia), willow (Salix spp.) and deerbrush (Ceanothus integerrimus). While the western marsh and shallow waters around the lake support an array of emergent, floating, and submerged aquatic plants. Common emergent species include three-way sedge (Dulichium arundinaceum), bogbean (Menyanthes trifoliata), tule (Schoenoplectus acutus) and Carex sedges; common aquatic plants include watershield (Brasenia schreberi), floating pondweed (Potamogeton natans), yellow pond lily (Nuphar lutea), and swaying bulrush (Schoenoplectus subterminalis).

The Mediterranean climate of the western Sierra Nevada is defined by warm, dry summers and cool, wet winters, and is governed by seasonal atmospheric circulation patterns in the North Pacific. During the winter months, the Aleutian Low strengthens and the Pacific Subtropical High moves equatorward, displacing the westerlies over northern California and allowing storms originating in the uppermost radiocarbon ages in each core (Table 1) and a co-occurring tephra (57 cm in Core 8a) (Starratt and Anderson, in press). Sampling occurred at ~5 cm intervals throughout the core, but more frequently (1–3 cm) over the upper 130 cm. A single sample was also collected from the topmost sediments of another freeze core (06-05, Roach, 2010), which recovered sediments from the last ~30 years that were missing from Core 8a (Anderson, 2011). Cores 8a and 06-05 were correlated based on measured 137Cs peaks (1964 C.E.) and several charcoal layers traceable to 20th century fires in the vicinity of SL.

### 2.2. Sediment cores

A series of sediment cores was recovered from Swamp Lake in June 2002, as described by Anderson (2011). In this study we used a 9-drive, 947 cm Livingstone core (Core 5) and a frozen finger (Core 8a) of the upper 83 cm, collected at a depth of 19.7 m (Fig. 1c). Cores 5 and 8a were composited based on the uppermost radiocarbon ages in each core (Table 1) and a co-occurring tephra (57 cm in Core 8a) (Starratt and Anderson, in press). Sampling occurred at ~5 cm intervals throughout the core, but more frequently (1–3 cm) over the upper 130 cm. A single sample was also collected from the topmost sediments of another freeze core (06-05, Roach, 2010), which recovered sediments from the last ~30 years that were missing from Core 8a (Anderson, 2011). Cores 8a and 06-05 were correlated based on measured 137Cs peaks (1964 C.E.) and several charcoal layers traceable to 20th century fires in the vicinity of SL.

### 2.2.1. Age control

Age control for this study is provided by thirteen AMS radiocarbon dates on bulk sediment OM (Table 1), a tephra geochemically identified as the Mazama ash at 400 cm (7.6 kyr BP) (Hallett et al., 1997; Anderson, 2011), and in the upper sediments of the freeze cores, measurements of 137Cs (peak at 1964 C.E.) and a charcoal layer representing a large local fire in 1996, verifiable by counting of visible varves (Roach, 2010; Anderson, 2011). Radiocarbon ages were converted to calibrated ages before 1950 C.E. (cal. yrs BP) using Calib 5.0.2 and the IntCal04 calibration dataset (Reimer et al., 2004; Stuiver et al. 2005). An age–depth model was developed using linear interpolation between calendar ages and 4.1 °C, respectively (1955–2008). Annual precipitation averaged 122 ± 41 cm, with ~90% occurring between November and May. Annual snowfall averaged 293 ± 109 cm (WRCC, 2012), representing perhaps 25–40% of total annual precipitation.

### Table 1

AMS radiocarbon age data for Swamp Lake cores 02-5 and 02-8a. All ages measured on bulk sediment.

<table>
<thead>
<tr>
<th>Lab ID</th>
<th>Core</th>
<th>Composite core depth (cm)</th>
<th>14C age (yr BP)</th>
<th>Calendar age (cal. yr BP)</th>
<th>Error (±2σ) (cal. yr)</th>
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<td>2789</td>
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(radiometric, tephra, charcoal) (Fig. 3). As shown in Table 1, 2σ errors for the calibrated ages range from ±36 to ±304 years, with greater age uncertainty in the lower part of the core. All ages given in the text refer to calibrated, calendar ages. Below the oldest radiocarbon date at 865 cm the age–depth relationship is unconstrained; dates from this interval (865–947 cm) should be treated with caution. The small size of the SL catchment and the absence of carbonates in the bedrock and lake sediments minimize the risk of age errors associated with the use of bulk sediment for radiocarbon analysis.

2.3. Field sampling

To determine the C and N isotopic values and C/N ratios of potential sources of OM to SL sediments, samples of the most common terrestrial and aquatic plant species at SL were collected during the growing season (May–Oct., 2007–2010). Suspended particulate OM (POM) was also collected at multiple depths on pre-combusted Whatman GF/F filters. Plant and POM samples were stored on ice in the field and frozen upon returning to the lab. Depth profiles of lake water temperature were taken in the central basin using a YSI 85/50 hand-held probe (Fig. 2a). Water samples (100–300 mL) for chl a analysis were collected along depth transects (Fig. 2b), filtered through GF/F filters and extracted in vials containing 90% acetone. The vials were stored on ice in the dark in the field and placed in a 4 °C refrigerator after returning to the lab. After 24 h, the samples were analyzed on a Turner fluorometer before and after acidification for phaeopigment correction.

2.4. Sample analysis

Sediment, plant, and POM samples were dried, homogenized and weighed into tin capsules for analysis. Total organic carbon (TOC), total nitrogen (TN), δ¹³C орг, and δ¹⁵N were determined using a PDZ Europa ANCA-GSL elemental analyzer interfaced to a PDZ Europa 20-20 continuous flow isotope ratio mass spectrometer (Sercon Ltd., Cheshire, UK) in the dual-isotope mode. Carbon and nitrogen isotope ratios are expressed in conventional delta (δ) notation as the per mil deviation (‰) from the Vienna PeeDee Belemnite (VPDB) carbonate rock standard and from atmospheric N₂, respectively. Instrumental precision, based on laboratory standards and measured as the standard deviation of all standard analyses (n = 115) was ±0.05‰ for δ¹³C орг and ±0.2‰ for δ¹⁵N. The reproducibility of TOC and TN measurements, estimated as the pooled standard deviation of duplicate analyses (n = 66, 0.1–21 wt % C), was ±6.8% (avg. ±0.4 wt% C) and ±9.3% (avg. ±0.04 wt% N), respectively. Though carbonates were not expected to be present in SL sediments, a subset of sediment samples were fumigated in a chamber with concentrated HCl (Harris et al., 2001) and compared with unacidified duplicates. TOC and δ¹³C values were unaffected by acidification, which we took as evidence that carbonates were absent or of negligible abundance in these sediments. Biogenic silica (BSI) was extracted from sediments with 10% Na₂CO₃ and concentrations determined with a spectrophotometer (Hach DR/2000) with an estimated precision of ±9 mg/g (±6%), following Mortlock and Froelich (1989).

3. Results

3.1. Geochemical and isotopic signatures of vegetation & POM

C/N and δ¹³C measurements of vegetation and POM samples collected in the summers of 2007–2010 allow for the mapping of distinct chemical signatures for potential OM sources in the modern lake basin (Fig. 4). Terrestrial vegetation was characterized by high C/N ratios (40.1 ± 24.6 (1σ), n = 15) and intermediate δ¹³C орг values (−28.8 ± 1.6‰), submerged/floating aquatic vegetation had intermediate C/N ratios (16.4 ± 2.9, n = 8) and higher δ¹³C орг values (−24.9 ± 1.9‰), with emergent aquatic plants plotting in between (C/N = 25.2 ± 10.2; δ¹³C орг = −26.9 ± 10.0‰; n = 7). Lake POM had low C/N (9.7 ± 2.9, n = 29) and δ¹³C орг (−34.6 ± 2.8‰) values. The C/N and δ¹³C орг values of these OM source groups were statistically distinguishable from one another above the 99% confidence level (p ≤ 0.01, Mann–Whitney U-test) with the exception of emergent and terrestrial plants. POM samples from the hypolimnion (≤6 m) had very low δ¹³C values (−37.2 ± 1.9‰; n = 10), distinct from epilimnion (−32.8 ± 1.9‰, n = 12) and thermocline (−33.2 ± 2.0‰, n = 6) samples (Fig. 4). The δ¹⁵N values of potential OM sources were less unique and more variable than their C/N and δ¹³C values, with no clear distinction between lake POM and vascular plants. Terrestrial, emergent, and submerged/floating aquatic plants had δ¹⁵N values of −2.6 ± 2.9‰, 0 ± 1.3‰ and −3.2 ± 1.8‰, respectively. The δ¹⁵N of lake POM (−0.2 ± 3.6‰) ranged from −12.1 to +4.0‰.

3.2. Sediment lithology

The SL composite core is dominated by fine sediment units, distinguishable from one another by OM content (low OM clays/silts vs. high OM gyttjas) and structure (laminated vs. massive) (Figs. 5 and 6a). The late Pleistocene portion of the record (947–560 cm; 19.7–10.7 kyr BP) consists largely of alternating beds of clay/silt and gyttja (massive or laminated), with multiple sand lenses and tephra. More significant sand layers occur at 658–642 cm (13.2–12.9 kyr BP) and 581–572 cm (11.0–10.8 kyr BP). A high OM peat layer occurs at 624–616 cm (12.6–12.1 kyr BP). The Holocene section of the core (559–0 cm; 10.7–0 kyr BP) consists of laminated gyttja, with occasional narrow bands of massive gyttja, clay, or sand. A peat layer occurs from 411 to 401 cm (7.8–7.6 kyr BP). A thicker interval of massive gyttja occurs in sediments sampled by the freeze core (48–19 cm; 425–112 yr BP). Thin-section analysis indicates that the
Laminations present throughout the SL cores represent annual varves (Roach, 2010; Anderson, 2011) consisting of light–dark couplets. Light layers are dominated by diatom frustules deposited during warm-season blooms, while dark layers are a mixture of OM, charcoal and clastic fragments associated with wet-season runoff into the lake (Roach, 2010).

3.3. Down-core variations in TOC, TN, C/N, $\delta^{13}$C$_{org}$, $\delta^{15}$N, BSi and lithology

Large changes in the amount and composition of OM accumulating in SL sediments have occurred over the past 20,000 years, signaling large rearrangements of the surrounding environment (Fig. 6b–f). Organic content (TOC, TN) increased 100-fold from the base of the core (0.27 wt% C, 0.1 wt% N) to the present (27 wt% C, 2.8 wt% N). Patterns of TOC and TN accumulation were nearly identical throughout the core ($r = 0.97$, $p < 0.0001$), and henceforth we generally discuss only changes in TOC. Superimposed on the long-term trend was a great deal of variability on multiple timescales, which is also reflected in the C/N ratio (6.9–21.1), $\delta^{13}$C$_{org}$ ($-22.2_{\text{min}}$ to $-35.5_{\text{max}}$), $\delta^{15}$N ($-2.8_{\text{min}}$ to $2.0_{\text{max}}$) and BSi content (5–556 mg/g) of sediments. The long-term average C/N ratio increased between 19.7 and 15.0 kyr BP, and declined toward the present. Long-term average $\delta^{13}$C$_{org}$ values declined between 19.7 and 14.0 kyr BP, increased from 14.0 to 10.8 kyr BP, and declined through most of the Holocene.

Fluctuations in TOC, C/N, $\delta^{13}$C$_{org}$ and $\delta^{15}$N were significantly associated with one another (Fig. 4, Fig. 7a–e; Table 2) through most of the SL record. TOC was negatively related to C/N, $\delta^{13}$C$_{org}$, and $\delta^{15}$N, which in turn were positively correlated (Table 2). Deviations from these relationships occurred during certain intervals, most notably the peat at 411–401 cm (7.8–7.6 kyr BP), several extreme low TOC intervals of the late Pleistocene (w 19.7–18.2, 17.3, 14.5, 16.4–16.2,10.8 kyr BP), and the high TOC interval between 751 and 721 cm (15.8–15.0 kyr BP, SLZ-4). TOC covaried with BSi (Fig. 6g) throughout the record, but were not compared quantitatively because the analyses were not done on the same samples.

3.3.1. Late Pleistocene (19.7–10.7 kyr BP)

The Late Pleistocene (LP) section of the core (947–560 cm, 19.7–10.7 kyr BP) contains nine alternating low- and high-TOC intervals (Fig. 6; “Swamp Lake Zones”, SLZ, 1–9) that generally
correspond to changes in sediment type. The earliest interval (SLZ-1, 947–878 cm, 19.7–18.2 kyr BP) is composed of gray clay and silt with low organic content (<1 wt% C, <0.12 wt% N), low C/N (7–12), high δ13Corg (−23.3 to −28.5\%o) and δ15N (0–1.4\%o) (Fig. 6a–f). Following SLZ-1 were eight alternating TOC maxima (centered at ~16.9, 15.3, 13.5, 11.1 kyr BP) and minima (~16.4, 14.5, 12.6, and 10.8 kyr BP).

All high TOC intervals consisted of massive and laminated gyttja (Fig. 6a), but nonetheless differed in important respects. During the earlier TOC maxima (SLZ-2b, 839–797 cm, 17.5–16.5 kyr BP; SLZ-4, 768–717 cm, 15.8–15.0 kyr BP), moderately high TOC (6–12 wt%) corresponded to distinct C/N maxima (14.6–21.1) and δ13Corg minima (~32.8 to ~30\%o). The latter TOC maxima (SLZ-6, 702–662 cm, 13.9–13.2 kyr BP; SLZ-8, 608–587 cm, 11.4–11.0 kyr BP) reached higher peak values (15.3 and 16.8 wt% C, respectively), but with C/N (14.7–20.9) and δ13Corg values (~32 to −25.6\%o) less distinct from neighboring low TOC intervals. The LP high TOC intervals generally corresponded to local BSI maxima (especially SLZ-4, 331–469 mg/g).

LP TOC minima (SLZ-3, 794–767 cm, 16.5–15.9 kyr BP; SLZ-5, 716–702 cm, 14.9–13.9 kyr BP; SLZ-7, 655–612 cm, 13.1–11.8 kyr BP; SLZ-9, 583–560 cm, 11.0–10.7 kyr BP) ranged from 0.4 to 3.2 wt% C at their lowest values, and contained a more varied array of sediment types than TOC maxima. SLZ-3 consisted of massive and layered gray clay, similar to that of SLZ-1, while SLZ-5 (laminated gyttja alternating with clay and sand layers), SLZ-7 (sand, gyttja, peat) and SLZ-9 (massive + laminated gyttja, interbedded sand) were more lithologically complex. SLZ-7 contained several distinct zonations: Coarse sand and gravel (655–642 cm); laminated gyttja (642–637.5 cm); massive gyttja with wood and sand (637.5–624 cm); and peat (624–616 cm). SLZ-3 and SLZ-5 were characterized by locally low (high) C/N (δ13Corg) values, while SLZ-7 and SLZ-9 were associated with high or intermediate C/N and δ13Corg values. TOC and BSI minima often corresponded (e.g., at ~16.3, 14.0, 12.8 and 10.9 kyr BP). Following SLZ-7 (after ~11.8 kyr BP), the dominant timescale of variability appears to have decreased; SLZ-8 and -9 could alternatively be described as a series of century-scale fluctuations leading into the early Holocene.

3.3.2. Holocene (10.7 kyr BP–present)

The Holocene section of the SL core consists of laminated gyttja with brief intervals of sand, clay, or massive gyttja (Figs. Sand 6a). Sedimentary OM content was typically higher than during the LP (5–27 wt% C, 0.3–2.8 wt% N) but quite variable on centennial to millennial timescales (Fig. 6b, c). TOC increased over the Holocene in three relatively abrupt steps at 10.8–10.5, 8.0–7.4, and 3.0–2.8 kyr BP, corresponding to increased BSI and lowered C/N, δ13Corg and δ15N (Fig. 6c–g). Based on these proxy shifts we defined early (EH, 10.7–8.1 kyr BP, ~12 wt% C), middle (MH, 7.5–3.1 kyr BP, ~16 wt% C) and late Holocene (LH, 3.0 kyr BP–present, ~17 wt% C) sub-intervals, during which TOC fluctuated around relatively stationary long-term means. TOC was strongly anti-correlated with both C/N (11.2–19.4) and δ13Corg (~−35.5\%o to −22.2\%o) throughout the Holocene; C/N and δ13Corg were positively correlated, and typically lower than LP values. δ15N decreased by >1.5\%o between 10.8 and 1.4 kyr BP, before increasing by a similar amount between 1.4 and 0.3 kyr BP. Higher frequency variability in δ15N, however, was muted in comparison to the other proxies.

EH sediments had the lowest TOC (9–17 wt% C) and highest baseline C/N (12.9–18.6), δ13Corg (~−31 to −29\%o) and δ15N (~−11 to 0.2\%o) values of the Holocene interval. Low BSI values (27–150 mg/
N values increased, while BSi and the frequency variability was prominent, with eight parallel, multi-
than during the EH.
records. (of a sand layer at EH ended abruptly with a events occurred at 10.0, 9.4 and 8.9 kyr BP (Correlations among SL OM proxies. All correlations are signi
minima (C/N, d15N, d13Corg) events were low relative to the EH, while BSi
15 wt% C), on average 3
C/N, C/N minimum) occurred at 7.5, 7.2, 6.6
century
TOC, C/N and d13Corg, d15N, d15N v. d13Corg. Intervals represented by distinct symbols: SLZ-1, 19.7–18.2 kyr BP (•); other low TOC, (○); SLZ-4, 15.8–15.0 kyr BP (+); SLZ-7, (□); peat at 412–402 cm, 7.8–7.6 kyr BP (×); tephra at 108–111 cm, ~1.5 kyr BP (□).

Fig. 7. Relationships among bulk OM proxies. (a) C/N v. TOC; (b) d13Corg v. TOC; (c) d15N v. TOC; (d) d15N v. C/N; (e) d15N v. d13Corg. Correlations among bulk OM proxies. (a) C/N v. TOC; (b) d13Corg v. TOC; (c) d15N v. TOC; (d) d15N v. C/N; (e) d15N v. d13Corg. Intervals represented by distinct symbols: SLZ-1, 19.7–18.2 kyr BP (•); other low TOC, (○); SLZ-4, 15.8–15.0 kyr BP (+); SLZ-7, (□); peat at 412–402 cm, 7.8–7.6 kyr BP (×); tephra at 108–111 cm, ~1.5 kyr BP (□).

Table 2
Correlations among SL OM proxies. All correlations are significant at the 99.9% level (p < 0.0001).

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<td>-0.69*</td>
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</tr>
<tr>
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</tr>
<tr>
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<td>-</td>
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<td>0.52</td>
</tr>
<tr>
<td>d15N</td>
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<tr>
<td>d15N</td>
<td>-</td>
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* Peat, tephra, “low TOC” and/or SLZ-4 intervals excluded.

4. Discussion

4.1. Controls on sedimentary OM abundance

The OM abundance in SL sediments at any given time reflects the balance of OM inputs (autochthonous & allochthonous) and losses (heterotrophic respiration), as well as dilution by other components, such as biogenic silica and inorganic detrital sediment. Each of these terms responds to climatic and environmental changes. For instance, the productivity and composition of aquatic organisms and terrestrial plant communities are sensitive to temperature, growing season length, precipitation patterns, and changes in lake level and mixing regimes. Likewise, the amount and OM content of detrital input to the lake, and thus the extent to which it dilutes (or enriches) sedimentary OM, will vary with climate-driven changes in soil stability, vegetation cover, and runoff in the drainage basin. The pathways and rates of OM degradation in

scale fluctuations in these records did not correspond closely to the TOC, C/N and d13Corg records.

A distinct transition into the LH occurred between 3.0 and 2.8 kyr BP, consisting of large increases in TOC and BSi, and parallel declines in d13Corg, C/N, and d15N. TOC remained high (10–27 wt % OC) from 2.8 kyr BP to the present, but underwent three multi-
century fluctuations (also seen in BSi, d13Corg and C/N), with TOC maxima (d13Corg minima) at 2.7–1.8, 1.0–0.5, and 0 kyr BP. The most dramatic LH changes occurred in the last 1000 years. From a broad TOC maximum (d13Corg, C/N minimum) during the medieval period (1000–550 yr BP), TOC declined (d13Corg, C/N increased) gradually until the mid-15th century (~420 yr BP), after which an abrupt drop in TOC (increase in C/N, d13Corg, d15N) coincided with deposition of massive gyttja. During the last 150 years, TOC and BSI increased three-fold, alongside large declines in C/N, d13Corg and d15N.
the water column and sediments are also strongly influenced by environmental factors including productivity, temperature and oxygen status. While the effects of detrital input on TOC can often be inferred from sediment lithology and other proxies (e.g., magnetic susceptibility), it is not trivial to distinguish between the effects of changing OM production, water column regeneration and post-depositional diageneis.

If OM losses related to post-depositional diageneis dominated the TOC record at SL, a monotonic down-core trend of decreasing TOC might be expected. While TOC and TN both decrease drastically moving down-core (Fig. 6b–c), this decrease was discontinuous, occurring abruptly at depths corresponding to times of major, independently-established environmental changes (see §4.4–4.5). Much of the variability in the proxy records occurs cyclically, which is also difficult to explain in terms of diageneis alone. In addition, TOC and TN are strongly correlated throughout the record, ruling out major selective losses (or retention) of one element over time.

The presence of varves through much of the SL record is indicative of an anaerobic depositional environment (Roach, 2010; Anderson, 2011), which would tend to increase OM burial efficiency and preservation (Gong and Hollander, 1997; Lehmann et al., 2002; Galman et al., 2008). Anaerobic respiration has been observed to impart a distinct geochemical signature to residual OM, with increased C/N and decreased δ13Corg and δ15N relative to primary OM (from preferential degradation of N-rich, 13C and 15N-enriched labile compounds) (Lehmann et al., 2002; Teranes and Bernasconi, 2005; Li et al., 2008). The strong positive correlations among C/N, δ13Corg and δ15N in the varved portions of the SL record are thus evidence that anaerobic diageneic processes do not dominate the proxy signals.

In light of these considerations, we conclude that the predominant controls on TOC abundance in SL sediments are (a) changes in OM input, reflecting the productivity of lake algae, bacteria and aquatic plants as well as allochthonous inputs from terrestrial plants, (b) OM reworking prior to or soon after deposition, and (c) dilution by inorganic components. Changes in OM sources and cycling within the lake can be traced using C/N, δ13Corg and δ15N (§4.2), while changes in sediment lithology help constrain detrital input. Our interpretation of the bulk OM record is supplemented by measurements of BSi concentration (Fig. 6g) and sediment magnetic susceptibility (MS, Fig. 6h; Anderson, 2011). BSI is commonly used as a proxy for the component of lake productivity contributed by diatoms or other siliceous algae (e.g., Colman et al., 1995; Hu et al., 2003). Patterns of TOC and BSI abundance are correlated over the SL record, supporting our contention that lake productivity is an important contributor to TOC, and indicating that dilution by BSI is not the major control on TOC. In granitic Sierra Nevada catchments, MS is dominated by the signal from magnetic minerals in lightly-weathered clastic debris (Konrad and Clark, 1998; Benson et al., 2002). We thus interpret MS as a proxy for detrital input and watershed stability, and as an indicator of dilution effects in the OM record. This interpretation is supported by the observation that peak MS values are associated with low OM clay, silt and sand intervals (Fig. 6).

4.2. Sources of OM to Swamp Lake sediments

4.2.1. Evidence from C/N ratios and δ13Corg

Higher plants, lake algae and bacteria produce OM with distinct C/N and δ13C signatures, and these proxies have been used extensively to determine sources of sedimentary OM (e.g., Meyers, 1994; Prah et al., 1994; Meyers and Takemura, 1997; Tenzer et al., 1997; Talbot and Laerdal, 2000; Hollander and Smith, 2001). C/N ratios of sedimentary OM primarily reflect the proportion of algal/microbial (low C/N) and vascular plant (high C/N) derived material. The original source signature can, however, be altered by the preferential degradation of labile compounds during anoxic decomposition (Teranes and Bernasconi, 2005; Li et al., 2008) or masked by the presence of inorganic N in very low OM sediments (Meyers, 2003). Differences in δ13Corg arise from isotopic fractionations associated with different carbon fixation pathways (e.g., C3 vs. C4 plants), the δ13C-content of inorganic carbon sources (e.g., atmospheric CO2, dissolved CO2, bicarbonate, methane), and a variety of degradational processes (LaZerte and Szalados, 1982; O’Leary, 1988; Hollander and Smith, 2001; Meyers, 2003; Li et al., 2008). The δ13C of lake organisms in particular is also sensitive to other environmental variables, including productivity and growth rates, water temperature, light availability, and long-term changes in atmospheric CO2 concentrations (e.g., Hollander and McKenzie, 1991; Laws et al., 1995; Hodell and Schelske, 1998; Gu et al., 1999).

As shown in Fig. 4, the C/N and δ13C values of modern OM sources to SL sediments include terrestrial and aquatic plants, and lake POM — define a mixing space encompassing the C/N and δ13Corg values of most down-core sediment samples. Based on this observation, we argue that sedimentary C/N and δ13Corg are to a large extent controlled by variable OM contributions from source groups similar to those found at SL at present. Other controls may have operated on these proxies, but the close correlations between δ13Corg and C/N throughout the record (Fig. 4) and the negative relationship between δ13Corg and TOC (Fig. 7b) suggest that the OM source signal predominates.

Accordingly, sediment samples with low C/N ratios and low δ13Corg values contain a greater proportion of OM derived from lake algae and/or bacteria, while high C/N and δ13Corg indicate a greater proportion of higher plant material. The main cluster of points in Fig. 4 traces a continuum between the POM and aquatic (emergent and submerged/ floating) plant end-members, suggesting that OM contributions from aquatic higher plants have generally exceeded terrestrial plant inputs. Three clusters of points diverge from the dominant trend: (a) low C/N, high δ13Corg values from SLZ-1 and extreme low TOC intervals of SLZ-2, 3, 5 and 9, probably reflecting algal production under C-limited, oligotrophic conditions (§4.3.3.1.); (b) high δ13Corg values from the peat layer at the EH–MH transition, likely reflecting OM inputs from aquatic plants (§4.5.2); and (c) high C/N, low δ13Corg values from LP intervals (especially SLZ-4), reflecting the greater relative importance of OM inputs from terrestrial plants during these time periods (§4.4.1.1.).

4.2.2. Low δ13Corg values suggest a distinct hypolimnetic POM source

The δ13Corg values of modern POM (δ13Corg) (~30.2 to ~40.2‰; Figs. 2c and 4) and many down-core samples at SL (Fig. 4) are low in comparison to the expected range for lacustrine algae utilizing dissolved CO2 in equilibrium with the atmosphere (~24 to ~30‰; Meyers, 1994, 2003). Moreover, δ13CPO4 in the hypolimnion after the onset of stratification is significantly lower than epilimnetic δ13CPO4 (Figs. 2c and 4; §3.1). Low δ13C in POM and surface sediments are observed in lakes where algal productivity is supported to some degree by uptake of 13C-depleted, respired carbon — produced by OM regeneration in the water column or methano genesis (and subsequent methane oxidation) in anoxic sediments or bottom waters — or where 13C-depleted secondary bacterial biomass (e.g., methanotrophs) contributes substantially to the total OM pool (Rau, 1978; Meyers and Ishiwatari, 1993; Hollander and Smith, 2001; Baxtiken et al., 2003; Teranes and Bernasconi, 2005; Li et al., 2008).

The presence of varved, unbioturbated sediments through much of the SL record is indicative of persistent seasonal anoxia at the sediment surface, consistent with field measurements of dysoxia (DO < 3 mg/L) below 10 m depth in the modern lake. Depth profiles
of temperature, chl a and δ13Corg at SL (Fig. 2) hint at a seasonal progression during which summer stratification (Fig. 2a) and productivity at depth (≥10 m; Fig. 2b) promote microbial respiration and oxygen depletion in the hypolimnion and surface sediments. Under such conditions, anaerobic respiration of OM produces methane (δ13C ≈ −65 to −50‰) (Whiticar, 1999; Hollander and Smith, 2001), and thus a source of δ13C-depleted carbon for methane-oxidizing bacteria, which in turn produce δ13C-depleted CO₂ that may support algal productivity in the hypolimnion (deep chl a maximum, Fig. 2b). The δ13C-depleted DIC pool that accumulates in the hypolimnion in the summer is transferred to the epilimnion during winter mixing and leads to low δ13C of POM during the early spring bloom (Fig. 2c). The frequent occurrence of methanogenesis at SL is also suggested by the presence of bacterial hopanoid biomarkers with very low δ13C values (−90 to −40‰) throughout the sediment record (Street et al., unpubl. data). Low δ13C values in SL sediments therefore reflect (a) the effects of an anoxic lake bottom on the DIC pool and subsequent primary production, and (b) the direct input of δ13C-depleted bacterial biomass to the sediments.

In light of these considerations, we interpret the sedimentary δ13Corg record as a measure of the balance of allochthonous vs. autochthonous (including both algal and bacterial production) OM sources, as with C/N, but also as a gauge of deep-lake C cycling intensity and oxygen status.

4.2.3. Evidence from δ15N

Nitrogen isotope compositions (δ15N) of sedimentary OM have been used to identify OM and N sources in lakes (e.g., Hodell and Schelske, 1998; Brenner et al., 1999; Talbot and Laerdal, 2000; Herczeg et al., 2001), but application of this proxy is complicated by the complex biogeochemistry of N in lake systems (Meyers, 2003). In the SL record, positive correlations with δ13Corg and C/N (Fig. 7e–f; Table 2) suggest OM-source control of sedimentary δ15N, but the distribution of δ15N values in modern OM sources (§3.1) would result in an interpretation contradicting that suggested by δ13C and C/N (i.e., lower δ15N indicative of larger OM contribution from vascular plants). We suggest instead that the initial δ15N source signal has generally been altered by OM degradation and N cycling in the SL watershed and water column.

OM regeneration in the water column (§4.2.2) would be expected to generate δ15N-depleted ammonium (NH₄⁺), which, if incorporated into subsequent primary productivity, would yield biomass with low δ15N values (−4 to +1‰; Meyers and Ishiwatari, 1993; Li et al., 2008). Secondary production by heterotrophic bacteria in the water column or sediments may also contribute δ15N-depleted biomass to the sediments while altering the δ15N signature of residual primary OM (Lehmann et al., 2002). Modern δ15Norg averages ~0‰, but with a broad range and no clear spatial or temporal pattern, perhaps reflecting a complex interplay between N-utilization during primary production (increases δ15NPN), remineralization (decreases δ15NPN) and secondary microbial production. Low δ15N values (−3 to 0‰) in most down-core sediments nonetheless support the idea that OM recycling and microbial productivity influence the SL δ15N record.

High δ15N (~0‰) in the SL record is often associated with high C/N and δ13Corg (and low TOC) (Fig. 6, Fig. 7e–f), which we have interpreted as reflecting a greater proportion of vascular plant OM in the sediments. However, since high δ15N values are not consistently observed in modern plant samples (−70 to +16‰), we suspect that high δ15N in the sediment record instead reflects (a) reduced generation of δ15N-depleted DIN within the lake due to weakened OM regeneration, and/or (b) the contribution of partially-degraded, residual soil and marsh plant OM, which is 15N-enriched relative to fresh plant litter (Kendall, 1998). During the

SLZ-1 (and other extreme low TOC intervals), high δ15N was associated with high δ13Corg (Fig. 7f) but low C/N (Fig. 7e), which we attribute to algal OM production under oligotrophic, N-limited conditions (§4.3.3.1).

4.3. Environmental interpretation of the SL record

4.3.1. Sediment changes & long-term development of the SL basin

Sedimentological changes in the SL record (§3.2) reflect the changing influence of several factors, including landscape stability and runoff, lake level, and autochthonous production of biogenic material (OM, BSI). Clay deposition during SLZ-1 (Fig. 6) probably records high meltwater runoff and detrital input from alpine glaciers in the drainage (§4.3.2.1). We attribute the shift to organic sedimentation (gættja) after ~18.2 kyr BP (Fig. 6) to further glacial retreat, increased landscape stability and reduced runoff/sediment transport, and increased lake productivity in response to a warming climate. The long-term decline in clastic sedimentation over the LP (Fig. 5) probably reflects increasing vegetation density and reduced erosion in the SL drainage, consistent with a trend toward higher sedimentary C/N (to ~15 kyr BP, Fig. 6e), and pollen evidence for the gradual replacement of alpine steppe with conifer forest (Anderson, 2011). Higher frequency changes in sedimentation, discussed in detail in §4.4.1, reflect the interaction of changing runoff and detrital input patterns with in situ biogenic production. Within the gættja intervals that dominate the SL record, knowledge of the OM content, BSI and magnetic susceptibility are crucial for inferring the relative importance of detrital vs. autochthonous components, and thus formulating an environmental interpretation. Peat layers deposited between 12.5–12.2 and 7.8–7.6 kyr BP may represent intervals of low lake level.

4.3.2. Controls on TOC at SL

Through most of the SL record, sedimentary TOC is significantly correlated with other proxies (Fig. 7a–c; Table 2), allowing us to make inferences about the environmental conditions influencing TOC abundance. With a few exceptions, TOC is negatively related to C/N, δ13Corg, and δ15N, and positively related to BSI. High TOC is often associated with deposition of varved sediments (Fig. 6). We conclude that high TOC abundance at SL is associated with some combination of high lake productivity (and thus a high proportion of OM from algal or microbial sources), low dilution by detrital sediments, and lake bottom anoxia. Moreover, low δ13Corg values observed during many high TOC intervals (e.g., LH) suggest that OM regeneration, methanogenesis and microbial productivity in the hypolimnion and surface sediments are important factors contributing to high sedimentary TOC. While microbial activity may degrade a significant portion of primary OM during high TOC intervals, it also must support a large contingent of secondary biomass that is preserved in the sediments. The relative importance of ecosystem productivity versus detrital dilution in determining sedimentary TOC concentrations is difficult to disentangle, but based on the more frequent occurrence of clastic sediments and high MS values, dilution was probably a more significant factor during the LP interval.

4.3.3. Environmental end-members

Lake responses to climate and environmental change are mediated by multiple site-specific factors (e.g., size, depth, stratification and mixing regimes, ice cover, nutrient distribution, water inputs) and are thus complex and variable in space and time (Schindler, 1997). In the absence of long-term monitoring data from SL, proposed links between climate forcing and past changes in productivity and sedimentation remain speculative. Nonetheless, we argue that by examining proxy values and relationships during
intervals of the SL record when external environmental conditions are partially constrained, we can bolster our interpretive framework and reach a better understanding of the full record. Specifically, we contrast two periods, SLZ-1 (19.7–18.2 kyr BP) and the 20th century, as potential “end-member” states in a continuum of environmental conditions at SL.

4.3.3.1. SL at the LGM. SLZ-1 (19.7–18.5 kyr BP), though poorly constrained in time (2.21), represents the early SL after the onset of Tioga stage glacial retreat. Sedimentation consisted of low TOC, high MS detrital clay and silt (Fig. 6), likely reflecting input from a sparsely-vegetated post-glacial watershed. The combination of low C/N, high δ13Corg, and high δ15N in SLZ-1 sediments (Figs. 6 and 7) is consistent with low productivity in an oligotrophic (possibly turbid) lake with limited supplies of DIN and DIC, and minimal higher plant contributions of OM. It is also possible that inclusion of inorganic N has depressed the measured C/N ratio in these low OM sediments (Meyers, 2003). LGM climate in the Sierra Nevada was colder than present, particularly during winter, with significant downslope shifts in vegetation assemblages (Koehler and Anderson, 1994, 1995; Mock and Bartlein, 1995; Benson et al., 1998; Anderson, 2011). Cold water temperatures, persistent seasonal ice cover, low rates of OM regeneration, bacterial activity and low pCO2 (190 ppm vs. 280 ppm pre-industrial; Barnola et al., 2003) would have exacerbated C limitation and contributed to high δ13Corg values in sedimentary OM (Hollander and McKenzie, 1991; Laws et al., 1995; Gu et al., 1999; Li et al., 2008). The sedimentary characteristics of SLZ-1 may thus provide a “signature” for cold-climate, oligotrophic, and high runoff/detrital sediment-dominated conditions at SL.

4.3.3.2. SL in the 20th century. Proxy values in 20th century sediments from SL reflect a lake environment very different from that of the LGM (Fig. 6a–h). High concentrations of TOC (14–27 wt.%, Fig. 8a), TN (1.4–2.8 wt.%), and BSi (22–42 wt.%), along with low δ13Corg (−31.3 to −35.5‰), δ15N (−1 to −2.8‰), and C/N (11–13) in varved sediments are indicative of high rates of algal productivity, low detrital input, predominantly autochthonous OM sources, sediment anoxia, and vigorous OM regeneration and secondary microbial production. We also conclude, based on field measurements, that these ecosystem characteristics are associated with a monomictic, seasonally-stratified mixing regime (Fig. 2a), absent or intermittent seasonal ice cover (Roach, 2010), hypolimnetic oxygen depletion, and deep (>10 m) summer chl a and POM maxima (Fig. 2b).

The large changes observed in the proxy records during the last 150 years (Figs. 6 and 8) may also reflect major recent changes in the lake environment. Since 1850, TOC, TN and BSi in SL sediments have increased two- to four-fold, while C/N, δ13Corg and δ15N decreased markedly (by 4, 8‰, and 2.7‰, respectively), with the greatest changes generally occurring since ~1950. While it is conceivable that these trends are an artifact of OM diagenesis, the directions of change in the individual proxies, as well as the parallels between organic and inorganic (BSi) proxies, are better explained by changes in the lake environment (i.e., increased productivity, OM cycling).

Regional climate in the Sierra Nevada over the last ~100 years has been warm relative to the preceding Little Ice Age (LIA, ~1500–1850 CE; Graumlich, 1993; Clark and Gillespie, 1997; Lloyd and Graumlich, 1997; Konrad and Clark, 1998), and has experienced both summer and winter warming since 1950 in response to shifts in North Pacific atmospheric circulation and, possibly, global climate change (Dettinger and Cayan, 1995; Cayan et al., 2001; Higgins et al., 2002; Millar et al., 2004). Plotted in Fig. 8a is a time series of compositied Jan–Mar temperature for seven central California stations (Table 3; updated/expanded from Dettinger and Cayan, 1995), showing a warming of ~2 °C since 1950 that parallels the period of most drastic change in TOC at SL (Fig. 7a).

Fig. 8. Comparison of (a) Swamp Lake sedimentary TOC during the 20th century with: (b) Jan–Mar mean temperature anomalies composited from seven Sierra region climate stations (Table 1). For each station, anomalies were calculated as the deviation from the average Jan.–Mar. temperature over the period of record. Only the Sacramento station was active prior to 1930. The solid black line represents the 9-yr running mean for the composite dataset; gray lines represent 9-yr running means for the Hetch Hetchy and Cherry Valley stations nearest Swamp Lake: and (c) Water-year (Oct–Sep) precipitation % of average (1930–2010) for the San Joaquin Basin 5-Station Index (Cal DWR, http://cedc.water.ca.gov/). We extended this record to WY 1907 using a smaller number of available stations (1913–1929, 4 stations; 1911–12, 3 stations; 1907–1910, 2 stations). Solid line represents 9-yr running mean.

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warming and nutrient enrichment provide a plausible explanation for recent changes in the proxy records and by extension, other high TOC intervals in the SL record.

The connections between climate and OM abundance at SL are most likely indirect, involving feedbacks among lake mixing regimes, winter ice cover, and nutrient cycling in the lake system. Reduced winter ice cover, for example, has been observed to enhance total productivity in lakes by influencing the timing of lake overturning, the onset of stratification, and the length of the growing season (Strub et al., 1985; Goldman et al., 1989; Byron and Goldman, 1990; Schindler, 1997; Bleckner et al., 2002; Winder and Schindler, 2004; Tranvik et al., 2009). If winter warming eliminates ice cover altogether, hypolimnetic temperatures in the resulting monomictic lake will parallel the annual minimum epilimnetic temperature during the isothermal period, and will thus be more sensitive to long-term changes in winter air temperature than in a dimictic, seasonally ice-covered lake (Peeters et al., 2002). At SL, we speculate that reduced or eliminated ice cover and earlier snowmelt runoff during warm intervals results in a longer growing season, but also a longer stratified period, potentially magnifying the importance of processes occurring below the thermocline. Warmer hypolimnion temperatures and stronger/longer stratification would promote increased microbial activity, more rapid OM regeneration rates, oxygen depletion, and, perhaps most crucially, increased microbially-mediated phosphorus (P) release from anoxic sediments (Prevert, 1980; Carlton and Wetzel, 1988; Pace and Prairie, 2005; Hupfer and Lewandowski, 2008; Tranvik et al., 2009; Gudasz et al., 2010). At SL, increased supplies of regenerated DIC and DIN, combined with sediment-sourced P, would support higher rates of summer productivity in the photic hypolimnion, the biomass from which would further fuel the microbial ecosystem, anoxia and methanogenesis in deep waters and/or surficial sediments. An increased supply of regenerated and sediment-released nutrients from the hypolimnion would then be available to support larger epilimnetic spring blooms in subsequent years. Some similar set of mechanisms (plus an external, anthropogenic nutrient source) is consistent with trends toward lower $\delta^{13}$Corg, $\delta^{15}$N and C/N, and higher TOC and BSi, in 20th century sediments at SL. Moreover, the release of P from anoxic, high OM, microbially-active sediments provides a plausible pathway for nutrient enrichment during past intervals, absent anthropogenic interference.

4.4. Late Pleistocene environmental history

4.4.1. Millennial-scale variability

4.4.1.1. High TOC intervals (SLZ-2, -4, -6, -8). The LP record was dominated by the deposition of alternating high- and low-TOC sediment units (Fig. 6, SLZ-2 to SLZ-9). The four high-TOC, gyttja-dominated intervals of the LP (SLZ-2, 18.2–16.5 kyr BP; SLZ-4, 15.8–15.0 kyr BP; SLZ-6, 13.8–13.2 kyr BP; SLZ-8, 11.4–11.0 kyr BP) can be interpreted as periods of increased ecosystem productivity and reduced dilution of sedimentary TOC, reflecting drier, warmer climate conditions and reduced sediment transport relative to SLZ-1. The combined effects of warmer air and water temperatures, smaller seasonal snowpacks, and the absence of glacial meltwater and till input to the lake would have reduced winter ice cover, stimulated lake productivity, and promoted vegetation growth and soil stabilization in the SL drainage, with the net effect of increasing OM content and reducing detrital input to the sediments. This interpretation is consistent with the relatively high BSI and low MS measured in the sediments of these periods. Relatively low $\delta^{13}$Corg (−33 to −28‰) (lowest values correspond to TOC maxima) may be indicative of OM recycling and microbial production under low oxygen conditions in the hypolimnion of a warm, stratified lake, although laminated sediments occur only intermittently during these intervals (e.g., SLZ-2, SLZ-8).

The unique combination of moderate-to-high BSI, high C/N and low $\delta^{13}$Corg in sediments from SLZ-2b and SLZ-4 (Figs. 4, 6 and 7) suggests that OM derived primarily from a mixture of lake algal/ microbial and terrestrial plant sources, with reduced contributions from aquatic plants in comparison to later periods. The long-term trend toward lower C/N values after SLZ-4 (~15 kyr BP), including SLZ-6 and ~8, may suggest an increasing presence of aquatic vegetation.

4.4.1.2. Low TOC intervals (SLZ-3, -5, -7, -9). The four low TOC intervals of the LP (SLZ-3, 16.5–15.9 kyr BP; SLZ-5, 14.8–13.9 kyr BP; SLZ-7, 13.1–11.8 kyr BP; SLZ-9, 11.0–10.7 kyr BP) each reflect lake environments in which OM production was reduced relative to dilution, but lithological and proxy differences among these intervals suggest that the causes of low TOC varied. The TOC minimum of SLZ-3, consisting of massive gray clay, closely resembles SLZ-1 (i.e., very low TOC, BSI, and C/N; relatively high $\delta^{13}$Corg, $\delta^{15}$N, MS), and likely represents a similar environment: a cold, oligotrophic lake, with high detrital input due to sparse watershed vegetation, and, possibly, a glacial advance back into the drainage. Clay deposition recurred briefly during SLZ-5, but for the most part the SLZ-5, -7 and -9 intervals consist of low OM gyttja, including multiple laminated units. Low TOC during SLZ-5, -7, and -9 is attributable to the combined effects of high detrital input (high MS) and reduced lake productivity, related to high runoff (large winter snowpacks), cold temperatures, persistent seasonal ice cover, and low rates of OM recycling and microbial production in the deep lake (high $\delta^{13}$Corg). However, the sediment types (gyttja vs. clay), less extreme TOC minima (2–6 wt%) and intermediate C/N values (mixed plant & algal sources) of SLZ-5, -7 and -9 in comparison to SLZ-1 and -3 signal that the later low-TOC intervals retained well-vegetated forest ecosystems and did not involve glacial advances into the drainage. These inferences are supported by pollen evidence for a permanent increase in the abundance of sub-alpine conifers (and decline of alpine tundra species) near SL after ~16 kyr BP (Smith and Anderson, 1992; Anderson, 2011) and evidence that Sierra glaciers were restricted to high elevations (>3000 m) after 15–14 kyr BP (Clark and Gillespie, 1997).

4.4.1.3. The Younger Dryas. SLZ-7 (13.1–11.6 kyr BP) in the SL record coincides with the Younger Dryas (YD) climate reversal and provides evidence for the expression of this event in the California region. SLZ-7 consists of several distinct sediment units, suggesting a complex structure for the YD interval at SL. A major sand/gravel layer (658–642 cm, 13.1–12.8 kyr BP) was followed by low TOC (2–6 wt%) and massive gyttja (12.8–12.5 kyr BP), high TOC peak (8–10 wt%, 12.5–12.2 kyr BP), and interbedded clay/gyttja/sand (12.1–11.6 kyr BP) (Figs. 5 and 6). The gyttja interval closely resembles previous low TOC intervals (with low BSI, high C/N, $\delta^{13}$Corg, MS), reflecting low lake productivity and elevated detrital input under cool, high runoff conditions (Fig. 4.1.2.); the preceding sand layer may represent a series of extreme runoff events. The peat interval, in contrast, reflects low lake levels. High TOC and BSI, along with low MS and $\delta^{13}$Corg are consistent with low detrital input and/or increased productivity and OM recycling, and may reflect warmer, low-runoff conditions (Fig. 4.1.1.). Our interpretation of the early YD (13.1–12.5 kyr BP) at SL agrees with previous evidence for depressed lake productivity and cold, wet conditions in the Sierra Nevada region at the height of the YD (Benson et al., 1998, 2010; Mensing, 2001; Porinchu et al., 2003; Briggs et al., 2005; MacDonald et al., 2008; Oster et al., 2009). Proxy patterns at SL after ~12.5 kyr BP are also consistent with regional
and suggested a teleconnection driven by shifts in the mean latitude of the polar jet stream (PJS) and the Pacific winter storm track in response to the expansion and contraction of the Laurentide Ice Sheet and Arctic sea ice (Sewall and Sloan, 2004; Kim et al., 2008). However, the SL record provides new evidence that climate changes in the Sierra Nevada predated Greenland warming by 2000–3000 years (Fig. 9), and that factors other than ice sheet dynamics were important determinants of regional climate during this time.

The SL TOC record (Fig. 9a) resembles SST reconstructions from the central and northern California margin (ODP 1017, Seki et al., 2002; ODP 1019, Barron et al., 2003) (Fig. 1a; 9b), suggestive of persistent links between ocean conditions in the California Current system and Sierra Nevada climate during the LP. Within age uncertainties (up to ±630 yr in 1019 age model), each major TOC shift at SL can be matched to a SST shift, such that high (low) TOC corresponded to relatively warm (cold) SST off California. A connection between coastal SST and the inland Sierra Nevada may be direct (i.e., buffering of continental temperature by the adjacent ocean) (Herbert et al., 2001) or indirect, mediated by co-occurring changes in the ocean-atmospheric circulation of the North Pacific. Modern studies have linked winter warming and early snowmelt in the Sierra Nevada since 1950 to a long-term increase in the incidence of deep low-pressure systems over the central North Pacific (strengthened Aleutian Low), leading to a northward displacement of the westerlies over North America, a higher frequency of warmer air masses reaching California.

Fig. 9. Late Pleistocene record of (a) sedimentary TOC at Swamp Lake, in comparison with (b) alkenone-derived SST reconstructions from ODP sites 1017 (Seki et al., 2002) and 1019 (Barron et al., 2003) off the California coast, (c) speleothem calcite δ13C, a hydrologic proxy, from Moaning Cavern in the central Sierra foothills (Oster et al., 2009), (d) lake level reconstruction for Owens Lake, adapted from Bacon et al. (2006), and (e) δ18O in the Greenland GISP2 ice core (Grootes and Stuiver, 1997). Gray shading highlights local maxima in proxy values (a, b, c, e) and low reconstructed lake level at Owens Lk. (d). SL sediment zones discussed in the text are labeled (SLZ-1 to -9), and suggested correlations in the other records identified by a corresponding number. Age errors for the ODP 1019 and Moaning Cavern records are estimated as the mean error (calendar yrs) of all radiocarbon ages between 10 and 20 kyr BP, from Barron et al. (2003) and Oster et al. (2009), respectively.
during the winter (Jan–Mar), and, in parallel, warmer winter SSTs along the California margin (Dettinger and Cayan, 1995; Mantua et al., 1997; Robertson and Chil, 1999; Higgins et al., 2002). The co-occurring changes at SL and California margin sites during the LP suggests some analogous North Pacific response to outside forcing, affecting the partitioning of sub-tropical and polar air masses reaching the central Sierra Nevada during the winter season, altering both temperature and precipitation regimes. The inferred warming at SLZ-2 (∼18.2 kyr BP) is consistent with observations of pre-Bolling warming in the North Pacific (Herbert et al., 2001; Seki et al., 2002; Kiefer and Kienast, 2005; Hill et al., 2006; Winograd et al., 2006), and the idea that tropical Pacific forcing contributed to patterns of deglacial warming in the extratropics (e.g., Lea et al., 2000; Kiefer and Kienast, 2005; Winograd et al., 2006).

4.5. Holocene environmental history at Swamp Lake

Relative to the LP, the Holocene at SL (<10.7 kyr BP) was characterized by high lake productivity and/or low detrital dilution, and increasingly autochthonous (algal, microbial, macrophytic) OM sources, as indicated by high sedimentary TOC and BSi, low C/N, δ13Corg and MS, and the predominance of varved gyttja throughout the interval. Varved sediments are suggestive of anoxic conditions at the sediment–water interface, consistent with high lake productivity and enhanced microbial reworking of OM and secondary productivity (low/declining δ13Corg, δ15N). These conditions are consistent with a warmer, dryer mean climate relative to the LP. Previous studies have estimated LP-Holocene warming of 3–5 °C at Sierra Nevada sites (Smith and Anderson, 1992; Mensing, 2001; Porinchu et al., 2003) supporting our contention that sustained high TOC in Holocene sediments broadly reflects temperature regime shift. Though millennial-scale fluctuations of the sort that characterized the LP were notably muted, the proxy records contained several abrupt shifts (4.5.1) and a high degree of centennial-scale variability (4.5.2).

4.5.1. Holocene regime shifts

Average TOC in Holocene sediments increased in three steps (each >2 wt% C), after ∼10.8, 8.0 and 3.0 kyr BP (Fig. 10a), with intervening periods characterized by stable long-term means. At each juncture BSi also increased, while C/N and δ13Corg decreased, a pattern consistent with increasing lake productivity and autochthonous OM, declining detrital input, and, as outlined in 4.2.2., an increasing influence of OM regeneration and secondary heterotrophic production in the hypolimnion on overall TOC abundance. We use these TOC increases to divide the Holocene record into early (EH, 10.5–8.0 kyr BP), middle (MH, 7.5–3.1 kyr BP) and late Holocene (LH, 3.0–0 kyr BP) sub-intervals, and argue that they were related to changes in cold-season temperature and hydrologic regimes, influencing detrital input and OM production and cycling within the lake, with the net effect of increasing sedimentary OM abundance.

4.5.1.1. Early Holocene (EH). Under this interpretive framework, comparatively lower TOC and BSi, and higher average MS, C/N and δ13Corg in EH sediments are indicative of higher vascular plant OM and detrital input relative to subsequent periods, potentially attributable to cooler, wetter conditions. However, previous studies at Sierra sites describe the EH as a period of warm summers, low

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**Fig. 10.** Holocene records of sedimentary (a) TOC and (b) δ13Corg at Swamp Lake, in comparison with alkenone-derived SST reconstructions from (c) ODP site 1019 (Barron et al., 2003) and (d) ODP site 1017 (Seki et al., 2002) off Northern California. Early (EH), middle (MH) and late (LH) transitions are indicated. TOC maxima (δ13Corg minima) are highlighted in gray; low TOC, high δ13Corg events are indicated by letters (a–s). Suggested corresponding events in the SL and ODP 1019 records are indicated by lettering. Age error for the ODP 1019 record is estimated as the mean error (cal. yrs) of all radiocarbon ages between 0 and 10 kyr BP (Barron et al., 2003).
effective moisture and high fire frequency, supporting open forests and extensive dryland/chaparral vegetation, associated with the northern hemisphere summer insolation maximum (winter minimum) at \( \sim 10 \) kyr BP (e.g., Anderson, 1990; Smith and Anderson, 1992; Anderson and Smith, 1994; Koehler and Anderson, 1995; Mensing, 2001; Brunelle and Anderson, 2003; Oster et al., 2009). A transition to warm-summer, seasonally-dry conditions in the SL drainage after \( \sim 10.8 \) kyr BP is also supported by pollen evidence for the arrival of alder and near-elimination of giant sequoia, fir and mountain hemlock (Anderson, 2011). If, as we have argued, OM production and cycling at SL are particularly sensitive to winter conditions (\( \leq 4.3k \)), the observed EH proxy values may reflect the combined effects of a highly seasonal climate: Cold winters, with increased ice cover and a delayed phytoplankton growing season, suppressed lake productivity and abetted detrital transport during spring runoff.

4.5.1.2. Middle Holocene (MH). The shift to higher baseline TOC after \( \sim 8.0 \) kyr BP was accompanied by increased BSI, lower C/N and \( \delta^{13}C_{\text{org}} \), and a permanent lowering of MS in SL sediments, together indicative of increased lake productivity, decreased detrital input, and an expansion of OM recycling and heterotrophic productivity. These patterns are consistent with a shift toward a drier climate and warmer winters relative to the EH. Evidence for a regional environmental transition \( \sim 8.0 \) kyr BP has been found in declining lake levels in basins draining the Sierra (Davis, 1999a; Benson et al., 2002; Bacon et al., 2006; Negrini et al., 2006) and the expansion/persistence of drought-tolerant vegetation on either side of the range (e.g., Anderson, 1990; Smith and Anderson, 1992; Davis, 1999a; Brunelle and Anderson, 2003; Mensing et al., 2004; Anderson, 2011). At SL, the period \( \approx 7.4–5.5 \) kyr BP appears as a broad BSI maximum (C/N minimum), indicative of increased algal (diatom) contributions to the OM pool relative to vascular plants. We speculate that these patterns represent an extended period of reduced runoff during the driest interval of the MH (\( \sim 7.5–6.0 \) kyr BP; Lindstrom, 1990; Benson et al., 2002; Mensing et al., 2004), when the lake was often restricted to its deep central basin, limiting the areal extent of the shallow water habitats utilized by aquatic/emergent vegetation (Fig. 1c). Between \( \approx 5.5 \) and 3.2 kyr BP, BSI declined and C/N increased gradually, reflecting reduced diatom productivity and an increased contribution of vascular plant material to sedimentary OM in response to increasing precipitation, runoff and lake levels, as has been inferred from other Sierra reconstructions (Smith and Anderson, 1992; Benson et al., 2002; Brunelle and Anderson, 2003; Mensing et al., 2004; Osleger et al., 2009).

4.5.1.3. Late Holocene (LH). After \( \sim 3.1 \) kyr BP, a major environmental shift occurred that over a few centuries led to large increases in TOC and BSI and declines in C/N, \( \delta^{13}C_{\text{org}} \) and \( \delta^{15}N \) in SL sediments. Since detrital input remained low (low MS) across the MH–LH boundary, these patterns represent a change in the trophic status of the lake, with increased lake primary productivity and expanded OM recycling and secondary production. The period of highest ecosystem productivity (high TOC, BSI) persisted until \( \sim 1.8 \) kyr BP, followed by a lower baseline lasting until the LIA interval (\( \sim 0.5 \) kyr BP). The MH–LH transition roughly corresponded to a series of regional environmental changes between \( \approx 0.4 \) and 3.0 kyr BP. These included glacial advances (Konrad and Clark, 1998) and treeline retreat in the southern Sierra Nevada (Scudder, 1987) and White Mountains (La Marche, 1973). Lake level increases and high stands in Sierra drainages (Stine, 1990; Davis, 1999a, b; Benson et al., 2002; Mensing et al., 2004; Bacon et al., 2006; Negrini et al., 2006), and increased effective moisture in western Sierra meadows and forests (Anderson, 1990; Smith and Anderson, 1992; Anderson and Smith, 1994; Brunelle and Anderson, 2003), consistent with a regional shift toward cooler summers and wetter winters. The evidence from SL suggests that increasing winter temperatures may also have been an important part of this environmental transition, inducing changes in winter ice cover, lake mixing regime, lake productivity and OM recycling via feedbacks similar to those outlined in \( \leq 4.3k \). Meanwhile, higher mean lake levels in response to a higher precipitation/evaporation balance may have allowed for the development of the shallow-water and marsh habitats that characterize the modern lake (Fig. 1c; Street et al., 2008; Anderson, 2011).

4.5.2. Holocene centennial variability

Beginning \( \sim 11.3 \) kyr BP, centennial-scale fluctuations in the SL proxy records (events a–s, Figs. 6 and 10), occurring within the longer-term regimes discussed above, became more prevalent. Though generally smaller in magnitude than LP millennial-scale events, these fluctuations significantly exceed measurement error (\( \pm 6\% \) of signal for \( >10 \) wt% C), encompass several data points, and occur in multiple proxies (e.g. TOC, TN, \( \delta^{13}C_{\text{org}} \)), implying a common environmental cause. The apparent shift in the dominant timescale of variability may reflect a change in the relative rates of climate forcing in the Sierra Nevada, or an increase in the sensitivity of the SL system. Low TOC (low BSI, high C/N, high \( \delta^{13}C_{\text{org}} \)) excursions represent intervals of increased detrital input and/or reduced ecosystem productivity and OM recycling. High TOC events likely reflect relatively eutrophic conditions with reduced detrital input. The more pronounced fluctuations observed in \( \delta^{13}C_{\text{org}} \) compared to C/N or \( \delta^{15}N \) (Fig. 6 and 10) suggest that the variability is linked to hypolimnetic OM production and carbon cycling. Multiple low TOC events during the EH and MH corresponded to coarse or massive sediment intervals, or small MS peaks (\( \approx 10 \) kyr BP; \( \approx 9.4 \) kyr BP; \( \approx 8.9 \) kyr BP; \( \approx 7.5 \) kyr BP; \( \approx 5.5 \) kyr BP; \( \approx 4.9 \) kyr BP; \( \approx 3.7 \) kyr BP; \( \approx 3.0 \) kyr BP) possibly representing high-runoff events and/or alleviation of lake bottom anoxia. Reduced vegetation cover or lower lake levels relative to the present may have facilitated sediment transport to the lake depocenter during such events.

The large excursion at the EH–MH transition (\( 7.8–7.6 \) kyr BP; Fig. 6 and 10) consists of a peat layer with high TOC, C/N, \( \delta^{13}C_{\text{org}} \), and \( \delta^{15}N \) values — a combination unique in the SL Holocene record. We speculate that the onset of very dry conditions at the EH–MH transition resulted in a drastic lowering of lake level, allowing aquatic plants (high C/N, \( \delta^{13}C_{\text{org}} \), Fig. 4) to colonize the lake shallows at high densities, in close proximity to the depocenter. High abundances of floating aquatic plants can be found at present in the shallow western end of SL, and covering several small, shallow lakes nearby.

The last thousand years in the SL record encompass large fluctuations in OM proxies that coincide with the relatively well-characterized Medieval and LIA intervals. High productivity and active OM cycling (high TOC, BSI, low \( \delta^{13}C_{\text{org}} \), C/N) at SL between \( \sim 1.0 \) and 0.5 kyr BP corresponded to generally warm and dry Medieval climate in the Sierra Nevada (e.g., Stine, 1990, 1994; Graumlich, 1993; Meko et al., 2001; Benson et al., 2002; Brunelle and Anderson, 2003; Mensing et al., 2004; Yuan et al., 2004; Hallett and Anderson, 2010). After \( \approx 0.55 \) kyr BP, declining TOC (BSI) and increasing C/N (\( \delta^{13}C_{\text{org}} \), \( \delta^{15}N \)) suggest a reduction in lake productivity and a relative increase in OM contributions from vascular plants, in conjunction with the onset of colder temperatures, glacial advances, treeline retreat, and reduced drought frequency in the Sierra during the LIA (e.g., Graumlich, 1993; Clark and Gillespie, 1997; Lloyd and Graumlich, 1997; Konrad and Clark, 1998). Deposition of massive gyttja between 0.4 and 0.1 kyr BP
(event “a”) may represent an intensification of these trends, and an alleviation of hypolimnetic anoxia, or a more rapid sedimentation event (Roach, 2010). SL diatom assemblages during the LIA interval are indicative of high lake levels and increased seasonal ice cover (Starratt and Anderson, in press). The consistent patterns in the proxy records over these recent, relatively well-constrained climatic intervals bolster our interpretation of Holocene centennial variability as primarily representing the response of the lake environment to alternating climate states.

4.5.3. Relation to California margin SST and North Pacific circulation during the Holocene

The major Holocene baseline increases in TOC after ∼10.8, 8.0 and 3.0 kyr BP can be matched to corresponding shifts in reconstructed California Current SST at Site 1019, within age uncertainties (Holocene mean ±270 yr for ODP 1019; ±36–188 yr for SL) (Fig. 10). Some century-scale fluctuations in TOC at SL may match specific even periods in the SST record (Fig. 10, events c, e, f, h, i, l, o, q, s). During the transitional period following the YD (∼12.0–10.0 kyr BP), high-amplitude SST fluctuations (≥3 °C) corresponded to the transitions between SLZ-7, -8, -9 and the EH in the TOC record. Following the SST/TOC peak at ∼10.0 kyr BP, a series of smaller, century-scale fluctuations in each record may also correspond (Fig. 10, events n–s). Warm reconstructed SST during the EH may be attributable to a mid-late summer weakening of the North Pacific High, slowing of the California current, and increased penetration of warm subtropical gyre waters into the coastal zone persisting into the fall–winter season (Barron et al., 2003; Barron and Bukry, 2007), perhaps suggesting a cold-season link between coastal SST patterns and TOC abundance patterns at SL.

Between 8.3 and 7.8 kyr BP (ODP 1019 age model), SST fluctuated markedly (∼1.3 °C), before settling 1–2 °C below the EH baseline for the next ∼4 kyr. This was interpreted as a general broadening of the California current and extension of the upwelling season, excluding gyre waters and reducing fall and winter SST (Barron et al., 2003; Barron and Bukry, 2007). Barron and Anderson (2011) attributed these oceanographic changes to the establishment of a North Pacific circulation regime analogous to the negative phase of the PDO, with a weak or west-trending winter Aleutian Low and a strong summer North Pacific High. Within the MH interval (∼8.0–3.0 ka), several century-scale TOC and SST fluctuations may correspond (Fig. 10) though the correlations become more tentative after ∼6.0 kyr BP.

Between 3.4 and 3.1 kyr BP (1019 age model), alkenone SST increased by ∼1.5 °C, attributable to increased fall and winter SST related to a seasonal weakening of the California current (Barron et al., 2003; Barron and Bukry, 2007). This regime shift, and the subsequent SST optimum (towards ∼2 kyr BP), are likely correlated with the large increase and sustained TOC increase in the SL record occurring after ∼3.0 kyr BP (3.0–1.8 kyr BP), which we attribute largely to higher winter–spring temperatures. LH oceanographic changes are consistent with a North Pacific regime-shift towards an El Nino/PDO-like state, characterized by a strong and/or eastward-shifted winter Aleutian Low, a weakened North Pacific High, and anomalously warm winter–spring temperatures and somewhat elevated precipitation in the Sierra Nevada (Zhang et al., 1997; Gershunov and Barnett, 1998; Cayan et al., 2001; Mantua and Hare, 2002; Barron et al., 2003; Castello and Shelton, 2004; Barron and Anderson, 2011). Warmer winter–spring conditions at SL after 3.0 kyr BP would likely have stimulated positive feedback loops in the lake ecosystem, ultimately resulting in the observed LH TOC increase (∼4.3; ∼4.5; ∼4.1).

In summary, most large changes in the abundance and composition of OM in SL sediments over the past 18,000 years were associated with SST shifts in the California current system off the northern and central California coast. This relationship was most noticeable during the millennial TOC fluctuations of the LP and major transitions of the Holocene, but may also be discernable during century-scale events of the Holocene. The relationship probably reflects large-scale changes in North Pacific ocean-atmospheric circulation, which altered temperature and precipitation regimes (especially during winter) in the Sierra Nevada in parallel with surface ocean conditions in the California current. The major exception to the generally positive SST–TOC relationship occurred at the EH–MH transition (∼8.0–7.5 kyr BP), when baseline TOC at SL increased while SST declined. Here, the local hydrologic effects of this climate transition (reduced runoff and detrital input, lower lake levels and changing mixing regimes) may have outweighed other influences. The abrupt increases in TOC at SL and SST off California at the MH–LH transition are broadly consistent with a late Holocene intensification of ENSO cycling in tropical Pacific records (e.g., Moy et al., 2002; Conroy et al., 2008) and a shift toward a ∼PDO-like mean state in the North Pacific circulation (Barron and Anderson, 2011).

5. Conclusions

The Swamp Lake OM record provides new insight into the patterns and pacing of environmental change in the Sierra Nevada since the LGM. Variability in sedimentary TOC abundance was closely associated with changes in OM composition (C/N, δ13Corg, δ15N), BSi concentration, total magnetic susceptibility, and sedimentology. These observations indicate that TOC over time has been governed by changes in lake total productivity (including both algal and heterotrophic sources), vascular plant OM inputs, and dilution by detrital materials eroded from the surrounding drainage. These factors in turn responded to changes in lake temperature, winter ice cover, mixing regimes, and spring–summer runoff in response to shifting climate conditions. Very low sedimentary δ13Corg values, particularly during high TOC, varved sediment intervals, combined with modern water column measurements (temperature, chl a, DO, δ13Corg) suggest that the intensity of OM regeneration and secondary microbial productivity under low-oxygen conditions in the hypolimnion and surficial sediments is an important control on sedimentary TOC.

The late Pleistocene (LP) record (19.7–10.7 kyr BP) was dominated by millennial-scale fluctuations between high and low TOC sedimentation, reflecting major changes in regional climate and local responses in the SL basin. Low TOC intervals (19.7–18.2, 16.5–15.8 kyr BP; 14.9–13.9 kyr BP; 13.1–11.7 kyr BP; 11.0–10.7 kyr BP), at times associated with increased deposition of clastic sediments, reflect oligotrophic conditions and heavy detrital input to the lake basin under cold, wet climates. Clay and silt deposition during earlier low TOC periods (19.7–18.2, 16.5–15.8 kyr BP) was related to erosion from poorly-vegetated, post-glacial landscapes; establishment of a sub-alpine forest ecosystem (by ∼15.0 kyr BP) resulted in a more stable watershed and less distinctive sedimentation during later low TOC intervals. High TOC gyttja intervals (17.4–16.5, 15.8–15.0, 13.9–13.2, 11.4–11.0 kyr BP), in contrast, are associated with reduced detrital dilution, increased lake productivity and plant OM inputs, and active deep-lake OM cycling under relatively warm/dry climate conditions.

After ∼12 kyr BP, SL sediments were predominantly varved, reflecting persistent seasonal anoxia in the deep lake and, after ∼10.5 kyr BP, sustained high productivity conditions. The Holocene record was defined by three abrupt increases in TOC after ∼10.8, 8.0 and 3.0 kyr BP, which we attribute to increases in lake productivity, deep lake OM cycling and secondary production and/or decreased detrital dilution and allochthonous OM input, related to...
warmer winter—spring temperatures and altered hydrologic regimes. Numerous centennial-scale fluctuations between warm/ dry (high TOC) and cool/wet (low TOC) conditions also occurred within the early, middle, and late Holocene regimes. Major changes in TOC and other proxies at SL on both millennial and centennial timescales corresponded to changing SST regimes in the California current system. The coherence between these records, persisting across all major climate transitions since the LGM, suggests a common forcing mechanism arising in the ocean-atmospheric circulation of the North Pacific.

Acknowledgments

Three anonymous referees contributed reviews that improved the quality of the paper. Scott Starratt, Douglas Hallett and John Barron also provided helpful comments at various points in this project. We thank Jan van Wagendonk, Dan Cayan, John Barron and James Stringfellow for logistical assistance; Daniel Boone, Jordan Bright, Alison Colwell, Elizabeth Derse, Douglas Hallett, Justin Holl, Erin Hult, Karen Knee, Tim Lambert, Jackie Liu, Ian McKenna, Nadine Quintana-Krupinski, Lydia Roach, and Benjamin Saenz for field assistance; and Cara Meeker and Susan Smith for laboratory assistance. Initial core characterization was carried out at LACore (National Lacustrine Core Facility), Department of Geology and Geophysics, University of Minnesota. Pyrolysis silica splits were made by Jordon Bright in the Sedimentary Records of Environmental Change laboratory at NAU. This research was supported by grants from the National Science Foundation (EAR-0902218 (A.P.), Calfed Bay-Delta Science Program (J.S.), and The Yosemite Fund (R.S.A.), and by funding from the U.S. Geological Survey. J.S. was supported by graduate student fellowships from Stanford University and the Calfed Bay-Delta Science Program. Laboratory of Paleocology Contribution # 132.

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